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The Earth's Rotation and Atmospheric Circulation—I Seasonal Variations

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Summary

Analysis of the observations of the variations in the Earth's speed of rotation reveals the usual semi-annual, annual and long-period terms. In addition these observations indicate the existence of a biennial term whose behaviour is intermittent. The geophysical and meteorological excitation functions have been evaluated and they are in very good agreement with the observed astronomical variations. The principal phenomena contributing to the total excitation function is the global zonal wind circulation up to altitudes of about 30 km. The irregular behaviour of the biennial term in the astronomical data enables some conclusions to be drawn about the biennial zonal wind characteristics. It appears that in 1960 and 1961 the downward propagation of the biennial winds was less pronounced than normal and that in the interval 1964 to 1968 the biennial wind component had either a period of about three years or its behaviour was confused at lower altitudes. From 1969 onwards the astronomical data indicates the existence of a strong biennial wind component propagating significantly down to altitudes of perhaps as low as 12 km.

1. Introduction

Annual variations in the Earth's rate of rotation, or in the length of day (l.o.d.) were reliably reported for the first time by Stoyko (1936). Some 20 years later Smith & Tucker (1953) reported the existence of a semi-annual term, part of which could be attributed to the bodily tide (Mintz & Munk 1951: Woolard 1959). This discovery led to a search for other tidal terms and by 1955 Markowitz (1955) reported both the monthly and fortnightly terms. The various results for the l.o.d. obtained up to about 1960 have been reviewed by Munk & MacDonald (1960) in their book *The Rotation of the Earth.*

Munk & MacDonald (1960) have also reviewed various meteorological and hydrological factors that could explain the variations in the length of day. The first correct interpretation—that these variations are caused by winds in the atmosphere or by variations in the angular momentum of the atmosphere—appears to have been made by Munk & Miller (1950). The question was investigated again by Mintz & Munk (1951, 1954) and in their latter paper they reached an order of magnitude agreement for the annual term but failed to explain the difference between the observed semi-annual term and the part due to the Earth tides. This failure appears to be due to the limited meteorological data available at that time, particularly data on the high altitude winds.

More recently Iijima & Okazaki (1966, 1971) have analysed the astronomical data and have reported a term with a period of about 26 months in the variations of the length of day, Δ(l.o.d.). They correctly interpreted this as being caused by the quasi-biennial component found relatively recently in the zonal winds, although their argument is qualitative only. Markowitz (1970) reports that this term has a period of 24 months rather than 26 months and that it is not strictly periodic: it dies out at times and later re-appears.

No serious attempt appears to have been made to re-evaluate the atmospheric excitation function since the publication of Munk & MacDonald's book, even though these authors stressed on several occasions that the problem is not solved and even though very substantial amounts of wind data have become available since 1960. We have re-opened the question in the anticipation of not only being able to adequately explain some of the observed variations in the Earth's rate of rotation but also to investigate the inverse problem; what can we deduce from the Earth's rotation in terms of the atmospheric circulation.

Apart from the seasonal variations in the length of day, long term variations also exist whose origin is usually relegated to the Earth's interior but there appears to be little agreement on how these variations can be explained (see for example, Rochester 1970; Runcorn 1968; Munk & MacDonald 1960). We have not attempted to interpret these 'non-seasonal' variations in this paper even though we have had to estimate them in order to isolate the seasonal variations. We have taken the point of view that before any variations are attributed to the inaccessible parts of the Earth, the influence of the observable atmosphere must first be fully explored.

2. Equations of motion

We use an orthogonal Earth fixed co-ordinate system whose x_3 axis lies close to the axis of rotation and whose x_1 axis lies near the Greenwich meridian. We introduce the following notation:

Ω = mean angular velocity of the Earth;

ω_3 = instantaneous angular velocity about the x_3 axis;

$m_3 = -\Delta(\text{l.o.d.})/\text{l.o.d.}$; $\omega_3 = \Omega(1+m_3)$;

I = maximum mean moment of inertia referred to the principal axis;

I_{33} = instantaneous moment of inertia referred to the x_3 axis;

$I_{33} = I+\Delta I_{33}$; $\Delta I_{33} = \Delta I_{33}(t)$;

h_3 = relative angular momentum about the x_3 axis due to the motion u_i of a volume element dV of density ρ;

$$h_3 = \int_V (x_1 u_2 - x_2 u)\, dV = h_3(t); \tag{1}$$

L_3 = exterior torque component along the x_3 axis.

With this notation the relevant equation of motion is (Munk & MacDonald 1960, page 38)

$$\left.\begin{aligned} m_3 &= \psi_3 \\ \psi_3 &= \frac{1}{\Omega^2 I}\left\{-\Omega^2\,\Delta I_{33} - \Omega h_3 + \Omega \int_0^t L_3\, dt\right\}. \end{aligned}\right\} \tag{2}$$

where ψ_3, the excitation function, contains all the geophysical phenomena influencing the astronomically observed variations in the length of day, m_3. The Earth's atmosphere can contribute to the excitation function (2) in three ways: (i) It can cause a change in the moment of inertia ΔI_{33}, and variations in atmospheric pressure contribute to the total variation in I_{33}. (ii) Winds acting along the Earth's surface exert torques along the x_i axes if the volume V is defined as that of the solid Earth; (iii) If the volume V is defined to include the atmosphere, the winds cause a change in the angular momenta h_i. Munk & MacDonald (1960, p. 128) have shown the formal equivalence of the torque and angular momentum approaches but they have also shown that the latter is a much more accurate method of computing the wind effects. Winds cause stresses or torques at the boundary over which they pass and these stresses are usually divided into two types; mountain stresses and surface stresses. The former are known from measurements to be very significant compared to the surface stresses but according to Smagorinsky (1969) simulations of atmospheric circulation with and without mountains give more or less the same total torques. This implies that the presence of the mountings only changes the relative partitioning of the two torques and it points to the extreme care that must be taken in using the torque approach. This has been demonstrated by Mintz (1951) whose calculations for the wind stresses caused by the Rocky Mountains alone yields almost twice the observed astronomical value for the Δ (l.o.d.).

Wind terms also occur in the equations for the polar motion. However, for geostrophic conditions, the total influence of the angular momentum variations on this motion must vanish (Munk & MacDonald 1960). Below 80 km the geostrophic and hydrostatic conditions are very reasonably satisfied (Murgatroyd 1969) so that the influence of winds on polar motion is insignificant: the surface and mountain stresses causing small perturbations in the polar motion only in so far as they are departures from the geostrophic condition.

The influence ΔI_{33} on the l.o.d. has been investigated by Munk & Hassan (1961) and they conclude that it is at most a few per cent of the angular momentum term. We therefore chose to ignore this term for the moment and discuss it in Section 5 together with other seasonal factors that contribute to the Δ(l.o.d.). The excitation function therefore reduces to (substituting (1) into (2)).

$$\psi_3 = \frac{-1}{\Omega I} \int_V \rho(x_1 u_2 - x_2 u_1)\, dV.$$

In terms of spherical co-ordinates r, ϕ, λ; $dV = r^2 \cos\phi\, dr d\phi d\lambda$, and

$$\psi_3 = \frac{-1}{\Omega I} \int_V \rho r^3 \cos^2 \phi u_\lambda{}^k\, dr d\phi d\lambda$$

where $u_\lambda = u_\lambda(r, \phi, \lambda)$ is the wind component normal to the meridian; that is, the zonal wind. The zonal wind data available is usually already averaged over all longitudes ($\bar{u}_\lambda$) and:

$$\psi_3 = -\frac{2\pi}{\Omega I} \int_{r=a_0}^{r+H} \int_{\phi=-\pi/2}^{\phi=\pi/2} \rho(r) r^3 \cos^2 \phi \bar{u}_\lambda(\phi, r)\, dr d\phi. \tag{3}$$

The $\cos^2 \phi$ under the integral suggests that the most important wind contribution to ψ_3 will come from equatorial and low latitude zones. The density $\rho(r)$ also indicates that high altitude winds will be less important even though they may reach maximum speeds there. The integral (3) furthermore indicates the caution that must be exercised when attempting to explain any variation in the Earth's rotation by regional meteorological phenomena such as the Asian Monsoon. These monsoons undoubtedly influence the Earth's rotation but only in so far as they are reflected in the global atmospheric circulation pattern.

3. Zonal wind circulation

The zonal wind pattern exhibits three principal periods; semi-annual, annual and biennial (or quasibiennial). Evidence for longer periods, such as an 11-yr cycle, is lacking (Dartt & Belmont 1970). The semi-annual zonal wind reaches a maximum amplitude of about 25 m s^{-1} near the stratopause in equatorial latitudes (Reed 1966; Quiroz & Miller 1967) and secondary maxima of about 5 m s^{-1} occur at about 12 km, also in equatorial latitudes (Newell *et al.* 1972). These maxima occur around the middle of April near the stratopause (Angell & Korshover 1970) and some weeks later at the lower level (Newell *et al.* 1972). The semi-annual oscillation extends to all latitudes as well as down to low altitudes. The annual zonal winds reach a maximum of about 70 m s^{-1} at about 60-km altitude at mid-latitudes, (Angell & Korshover 1970). Important secondary maxima of about 20 m s^{-1} occur at about 12 km also at mid-latitudes. The maximum values at 60-km altitude occur around the end of December and about a month later at the lower altitude. The quasibiennial zonal wind reaches an amplitude of about 20 m s^{-1} at an altitude of about 25 km in equatorial latitudes (Newell *et al.* 1972; Dartt & Belmont 1970; Angell & Korshover 1965, 1970) and is continuous up to altitudes of 60 km and more. The period of this zonal component has variously been reported as between 20 and 30 months (Dartt & Belmont 1964) although Wallace & Newell (1966) have suggested that it is a biennial term whose pattern is interrupted at intervals and later reactivated, possibly with a change of phase. The times at which the maxima occur at different altitudes and latitude is extremely variable. Results by Angell & Korshover (1970), for example, show that at polar and temperate latitudes there is nearly an out of phase relationship between the winds at 20- and 60-km altitudes.

There does not yet appear to be unanimous agreement whether or not the biennial zonal wind is an intermittent phenomenon or whether its period is 24 months rather than 26 or 28 months. Wallace & Newell (1966) indicated that a breakdown of the biennial period occurred in 1963 although the data of Angell & Korshover (1970) for the Northern Hemisphere suggests that the cycle is re-established by about 1965. The data of Angell & Korshover (1970) and of Dartt and Belmont (1970) indicate furthermore that there are rapid fluctuations with time in the altitudes at which the maximum west zonal winds occur. A biennial variation in atmospheric ozone has also been reported (Ramanathan 1963), and Kulkarni (1966) reports a breakdown in the biennial atmospheric temperature component after 1963. The biennial ozone and temperature variations have been connected with the biennial winds (Ramanathan 1963; Kulkarni 1963) and these observations essentially substantiate the suggestion of Wallace & Newell (1966). It is not the intention to discuss here the various explanations and theories associated with the biennial term. Rather, we have outlined the various pertinent factors in order to assist in the subsequent interpretation of the astronomic data and to determine to what extent the astronomic data can be used to verify certain hypotheses on the atmospheric circulation.

4. Zonal wind data and the excitation functions

The most complete global zonal wind compilation presently available is that by Newell *et al.* (1972) who give directly the amplitude and phase of the three principal zonal wind components as a function of latitude and altitude. The latitude range is between $\pm 40°$ and the altitude range is from 1000 to 20 mbar or up to about 27 km. The distribution of the observing stations is given by Kidson, Vincent & Newell (1969). The data used is essentially the same as that reported by Newell *et al.* (1969);

that is, from 1957 to 1964, except that data after July 1963 has been excluded as at this epoch the biennial component apparently vanished. The excitation functions corresponding to the three wind components are given by the first three entries in Table 1. Newell *et al.* (1969) also give the total mean zonal wind profiles for two three-month periods, December–February and June–August, for all latitudes. These data are useful for evaluating the influence of high latitude winds on the variations in the Earth's rotation rate, and provides, in consequence, an estimate of the error of omission resulting from the use of the Newell *et al.* (1972) data set. The fourth excitation function is for the total atmosphere and the fifth function is for the atmosphere between $\pm 40°$ latitude. Comparing the two functions shows that the higher latitudes contribute about 20 per cent to the total excitation function.

To obtain the excitation function for the atmosphere above 20 mb we made a harmonic analysis of the Belmont & Dartt (1970) data at 40 and 50 km for the phase and amplitude of the three principal wind components as a function of latitude and at the two altitudes. An interpolation in altitude was then made between these results and the phase–amplitude results given by Newell *et al.* (1972) to give the appropriate ψ_3. The ψ_3 are given in Table 1 entries 6, 7 and 8.

The annual zonal winds reach maximum speeds at higher altitudes and to obtain an order of magnitude of this contribution to the annual excitation function the mean wind data between about 60 km and 120 km given by Murgatroyd (1965) is used. His mean profiles are given at two epochs but because of assumptions of hemispherical symmetry the data cannot be used for estimating the true excitation function. The estimated amplitudes are given in entry 9 of Table 1. This contribution to ψ_3 is clearly unimportant; the decrease in atmospheric density outweighing the increased wind speeds.

In summary, the functions 1–6 appear to describe quite adequately the wind excitation functions although all will be underestimated because of the lack of high latitude wind data. The winds between 30 and 50 km contribute only marginally to the total excitation function even though the winds reach their maximum speeds here. This is due in part to the decrease in density with increasing altitude and in part to an increasing hemispherical symmetry of the annual zonal winds at high altitudes.

5. Additional seasonal excitation functions

As discussed in Section 2, variations in the moment of inertia about the rotation axis also cause variations in the Earth's rotation according to equation (2). One of the factors contributing to ΔI_{33} is the seasonal variation in pressure, as already

Table 1

Excitation function ψ_3 for the semi-annual, annual and biennial components in the zonal wind circulation. For the items 4, 5 *and* 9 *only the total amplitude of the ψ_3 could be computed. The time t is in months counted from* 1958 *January* 0. *The ψ_3 is expressed in the form A* $\cos(2\pi t/P) + B \sin(2\pi t/P)$

	Period P (months)	Altitude (km)	$A(10^{-8})$	$B(10^{-8})$	Data source
1	6	0–30	−0·051	0·260	Newell *et al.* (1972)
2	12	0–30	−0·344	−0·131	Newell *et al.* (1972)
3	24	0–30	−0·077	0·029	Newell *et al.* (1972)
4	Total winds	0–30	0·57		Newell *et al.* (1969)
5	Total winds	0–30	0·45		Newell *et al.* (1969)
6	6	30–60	0·050	0·010	Belmont & Dartt (1970)
7	12	30–60	−0·024	−0·010	Belmont & Dartt (1970)
8	24	30–60	−0·001	−0·003	Belmont & Dartt (1970)
9	Total winds	60–120	0·01		Murgatroyd (1965)

discussed in Section 2. Hassan (1961) has evaluated monthly mean values of ψ_3 due to variations in pressure for the years 1873 to 1950 and a harmonic analysis of these results gives amplitudes of $0{\cdot}48 \times 10^{-10}$ for the annual term, $0{\cdot}07 \times 10^{-10}$ for the semi-annual term, and a smaller term of $0{\cdot}03 \times 10^{-10}$, hardly above the noise level, at about 26 months, (Table 2).

Table 2

Excitation function due to phenomena other than the atmospheric circulation. Items 7 and 8 give the total semi-annual and annual excitation functions caused by atmospheric pressure variations, by variations in ground water storage and by the solid Earth tides.

	Period (months)	$\psi_3(10^{-8})$	Source
1	6	$-0{\cdot}0006 \cos \frac{2\pi t}{6} + 0{\cdot}0000 \sin \frac{2\pi t}{6}$	Modified excitation function derived from atmospheric pressure data of Hassan (1961)
2	12	$-0{\cdot}0032 \cos \frac{2\pi t}{12} + 0{\cdot}0027 \sin \frac{2\pi t}{12}$	
3	6	$0{\cdot}003 \cos \frac{2\pi t}{6} - 0{\cdot}005 \sin \frac{2\pi t}{6}$	Ground water (Hylckama 1970)
4	12	$0{\cdot}017 \cos \frac{2\pi t}{12} + 0{\cdot}022 \sin \frac{2\pi t}{12}$	
5	6	$0{\cdot}161 \cos \frac{2\pi t}{6} - 0{\cdot}060 \sin \frac{2\pi t}{6}$	Solid Earth tides
6	12	$-0{\cdot}028 \cos \frac{2\pi t}{12} - 0{\cdot}001 \sin \frac{2\pi t}{12}$	
7	6	$0{\cdot}162 \cos \frac{2\pi t}{6} - 0{\cdot}065 \sin \frac{2\pi t}{6}$	Sum of (1), (3) and (5)
8	12	$-0{\cdot}014 \cos \frac{2\pi t}{12} + 0{\cdot}024 \sin \frac{2\pi t}{12}$	Sum of (2), (4) and (6)

Under a surface load, the Earth deforms: there is a depression of the surface by the load but there is also an attraction of the Earth by the load itself. This deformation can be described by the load deformation coefficients k_2' (similarly defined as the Love number k_2) of degree 2. Seismic models for the Earth give $k_2' \approx -0{,}30$ (Longman 1966) and the effective excitation function due to surface loading becomes $(1+k_2')\psi_3$ for those functions that can be described as zonal harmonics of degree 2, that is all the seasonal variations. Table 2, entries 1 and 2, give the modified excitation functions for the annual and semi-annual terms.

The excitation functions due to variations in surface loads caused by ground water have been discussed by Munk & MacDonald (1960) and re-evaluated by Hylckama (1970) (entries 3 and 4, Table 2). They do not exceed a few parts in 10^{10}. Solid Earth tides are important, particularly for the semi-annual term. These excitation functions can be evaluated directly from the theory developed by Woolard (1959). Their amplitudes depend on the choice of the Love number k_2 and a nominal value of 0·29 applicable for static Earth models, will be correct to within 5 per cent (Jeffreys 1970). The appropriate excitation functions are given in Table 2, entries 5 and 6. Entries 7 and 8 give the total excitation functions due to all the geophysical sources other than winds. They are probably correct to a few parts in 10^{10}, the major uncertainty coming from the ground water excitation function. In any case, this uncertainty is small compared to the wind terms given in Table 1.

6. Astronomical data and its analysis

The astronomically observed quantities are not the m_3 discussed in Section 2 but the integrated amount by which the Earth is slow or fast with respect to a uniform time scale. Thus, if the time kept by the Earth is denoted by UT (Universal Time) and the reference time by R the measured quantity (by convention) is $\tau = -(\mathrm{UT}-\mathrm{R})$, τ being the amount by which the Earth is slow. The observed Universal Time is denoted by UT0 and gives the rotation about the x_3 axis since the observers meridian is defined by an Earth fixed reference system. The rotation about the instantaneous axis is denoted by UT1 and is the UT0 corrected for the variations in the directions between the x_i and the rotation axis. Universal Time corrected for nominal seasonal variations is denoted by UT2 but this is of no interest to us except that we have to remove this 'correction' from some of the data. Variations in the length of day (l.o.d.), τ and m_3 are related by the following expressions.

$$m_3 = -\frac{\Delta(\mathrm{l.o.d.})}{\mathrm{l.o.d.}} = -\frac{d\tau}{dt} = \frac{d(\mathrm{UT1}-\mathrm{R})}{dt}$$

m_3 therefore refers here to variations in rotation about the instantaneous axis.

Precise astronomical observations have been possible for many decades but reliable observations for τ became available only since 1955 when atomic clocks were introduced to provide the uniform reference scale R. Since 1955 the astronomical observations have been made from different stations with different instruments and have been reduced in different ways. Thus no truly homogeneous data set exists for the entire period 1955.5–1972.0 and we have adopted the following data set for our analysis.

(i) For the interval 1955.5–1962.0 the revised values of UT2–R published by the Bureau International de l'Heure (B.I.H.) (Guinot 1965) have been used. The following correction was made to reduce the data to the required UT1.

$$-\tau = \mathrm{UT1}-\mathrm{R} = (\mathrm{UT2}-\mathrm{R})-(\mathrm{UT2}-\mathrm{UT1})$$

with

$$\begin{aligned}\mathrm{UT2}-\mathrm{UT1} = {} & 22\,\mathrm{ms}\sin 2\pi t - 17\,\mathrm{ms}\cos 2\pi t \\ & -7\,\mathrm{ms}\sin 4\pi t + 6\,\mathrm{ms}\cos 4\pi t\end{aligned}$$

and where t is the fraction of the year.

For the interval 1955.5–1958.0 the atomic reference time, R was based on the unweighted mean of all atomic frequency standards communicated to the B.I.H. Subsequently the A3 time scale has been adopted and it is assumed that the transfer has been smooth.

(ii) For the interval 1962.0–1968.0 the revised but smoothed values published in the B.I.H. annual report of 1970 (Guinot, Feissel & Granveaud 1971) have been adopted. A small discontinuity occurs at 1962 January 0, due to the different reduction methods used before and after this date, but this is of no consequence for our analysis of relatively long periods.

(iii) For the interval 1968.0–1971.0 the smoothed values published in the corresponding annual reports of the B.I.H. are used. The recent data is with respect to the atomic time-scale AT which is a smooth continuation of A3.

(iv) For the year 1971. The data at five-day intervals published regularly in the circular D of the B.I.H. has been adopted. In contrast to the other data, no smoothing has been applied and in consequence this data shows a greater variability between successive points. This is again of no consequence for our analysis.

The accuracy of the data since 1962.0 is better that 0·002 s (or 2 ms) and for the period before 1962.0 the accuracy is probably better than about 3 ms. Thus any variation in the UT − R curve of amplitude greater than about 3 or 4 ms is physically significant and not due to observational errors. Short-period variations in the Earth's rotation occur due to the tidal deformations. Their amplitudes are about 1 ms for the 14-day and 28-day periods. The data we used show some evidence of these terms despite the smoothing that has been applied to the majority of the data but we have not attempted to determine them.

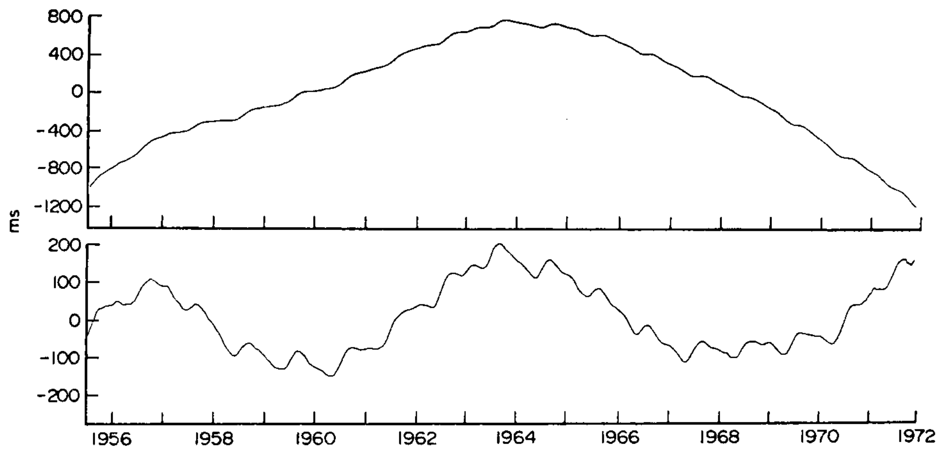

FIG. 1. Observed variations in $-\tau$; upper curve after elimination of a secular decrease and lower curve after elimination of a polynomial of degree 5.

The (UT − R) shows in the first instance the familiar secular decrease; a mean value of − 1·6 ms/day for the period 1955.6 to 1972.0. After eliminating this secular term the residuals show a parabolic trend (Fig. 1). Eliminating this, the residual differences are very suggestive of a periodic term whose period is about 7 or 7·5 years (Fig. 1). In order to study the seasonal variations, these long term variations must first be removed but there appears to be no common consensus on the best way to do this. Thus Markowitz (1970), following Brouwer (1952), uses a series of parabolas with common tangent points to represent the long-term variations; that is the speed of rotation varies linearly with time up to the tangent point at which instance a change in speed occurs. Stoyko & Stoyko (1966) and Iijima & Okazaki (1971) have used a polynomial plus periodic terms with periods equal to that of the 18.7 and 9·3 nutation terms. The 18·7-yr terms can be justified by the fact that there is an 18·7-yr period tidal-induced variation in UT1 − R with a theoretical amplitude of 0·5150 k_2s (Woolard 1959) or about 0·15 for $k_2s = 0{\cdot}30$. The 9·3 yr tidal term in the Earth rotation, however, has an amplitude of only 0·0027 k_2 s or about 0·001 s. The amplitude of the 18·6 term found by Iijima & Okazaki (1971) is 0·86 s or five or six times too large to be explained by the tidal contributions.

In our analysis we have only sought a simple and efficient method of eliminating the non-seasonal variations without attempting to give any physical significance to the terms, and a fifth degree polynomial was found most suitable. Higher degree terms did not reduce the residuals, but a significant periodic term of about 7·5 years remained, (Fig. 1). In the final least squares analysis the 5th degree polynomial and the long periodic term were determined simultaneously with the following results:

$$\mathrm{UT-R-P} = 1{\cdot}3876 - 1{\cdot}618\times10^{-3}\,t - 2{\cdot}316\times10^{-7}\,t^2 - 5{\cdot}961\times10^{-11}\,t^3 + 3{\cdot}360\times10^{-15}\,t^4 + 4{\cdot}217\times10^{-18}\,t^4$$

where t is in days since 1963 September 26, the middle of the observing interval. Co-variance analysis of the residuals UT − R − P verified that all variations of long period have been adequately eliminated. The variations shown in Fig. 2 clearly reveal the seasonal variations but they do show some irregularities. In particular, in the Northern Hemisphere spring of 1959, 1963 and 1968 the usual negative value is not as well developed as it is for the other years. These anomalies in 1959 and 1963 are followed by larger than usual maxima in September. Finally, the spring negative in 1970 is particularly poorly developed and preceded by a significant double positive anomaly in the latter part of 1970.

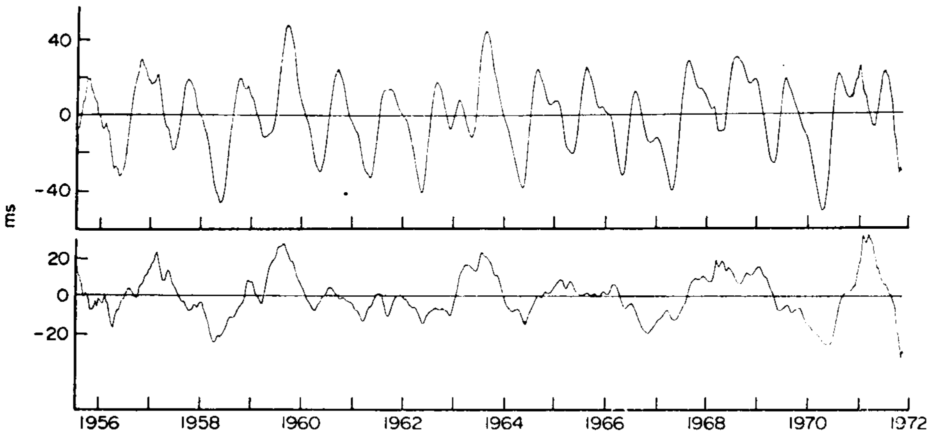

FIG. 2. Observed variation in $-\tau$ after the elimination of all long period terms (upper curve) and the residual variations in $-\tau$ after the elimination of the long-period terms and the semi-annual and annual variations.

Harmonic analysis of the UT − R − P clearly defines the annual and semi-annual terms (Fig. 3) and these can be represented by the expression

$$\mathrm{SV} = -10{\cdot}25\cos\frac{2\pi}{6}(t-5{\cdot}22) - 19{\cdot}81\cos\frac{2\pi}{12}(t-3{\cdot}80)$$

which is in good agreement with the results of Iijima & Okazaki (1971) and is as it should be since essentially the same data has been used. The harmonic analysis of the UT − R − P also reveals peaks at 15·5, 18·7, 24, 27·8 and 35 months. Some of these peaks are in general agreement with the biennial or quasibiennial zonal wind characteristics discussed in the previous sections, but the intermittent character of these winds suggests that we should not base our conclusions on the harmonic analyses but rather on the residuals UT − R − P − SV given in Fig. 2.

From these residuals we observe the following characteristics:

(i) From the beginning of the astronomical record in July 1955 to the spring of 1960 there is a clearly defined nearly biennial period with a period of 27 months and an amplitude of about 14 ms.

(ii) From April 1960 to January 1962 there is no evidence of a biennial term.

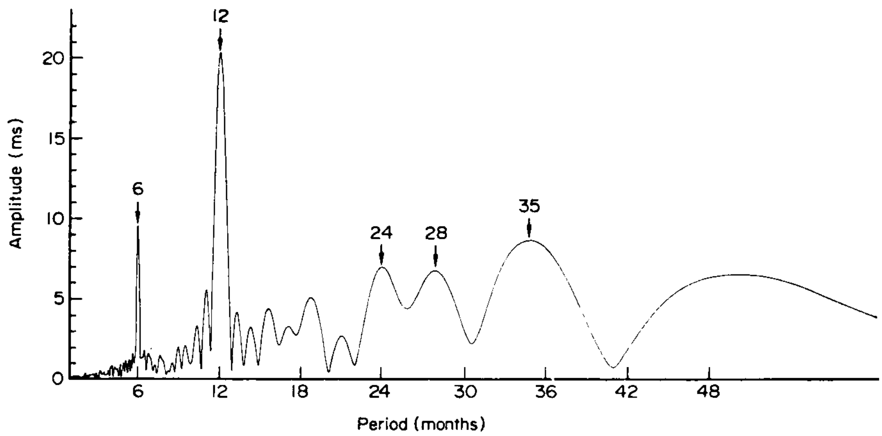

FIG. 3. Periodogram of the $-\tau$ after the elimination of the long-period variations.

(iii) From January 1962 to about January 1965 the biennial term is again present with an amplitude of 14 ms and a period of 24 months. The time at which the biennial term vanishes is not clearly defined and merges with a longer period term that appears in the next interval.

(iv) From early 1965 towards the end of 1969 there is no evidence of a biennial term but a longer period term, of 36 months, and 10 ms amplitude occurs.

(v) By January 1970, the biennial term has re-established itself with a period of 24 months and an amplitude of 24 ms. The date of its reappearance merges with the end of the long-period term of the previous time span.

(vi) Superimposed on the quasibiennial oscillations (Fig. 2) are residual annual and semi-annual terms due to the fact that these terms have variable amplitudes. Analyses of two year running mean indicates that the annual term's amplitude varies by not more than about 4 ms (or about 20 per cent) of its amplitude and by not more than about 5 or 6 weeks in phase. The amplitude of the semi-annual term is constant to within about 3 ms (or about 30 per cent) in amplitude and about 3 or 4 weeks in phase.

The variability of the amplitude and phase of the semi-annual and annual terms as well as the non-periodic behaviour of the biennial term indicate that the astronomical data should be analysed for the same period for which the ψ_3 has been evaluated in order to make a direct comparison of the m_3 and ψ_3. The τ and corresponding m_3 are given in Table 3. The astronomical data suggest a period of 26 months rather than 24 for the biennial term.

Table 3

Summary of the τ and m_3 estimated from the astronomical data from 1957 to 1963. The ψ_3 represents the total excitation function for the same period. The three functions are given in the form $A \cos (2\pi t/P)(t-\beta)$ where t is in months since 1958 January 0, and the phase β is in months.

Period			$m_3(10^{-8})$		$\psi_3(10^{-8})$	
P (months)	τ(ms)		A	β	A	β
6	8·65	5·27	0·346	0·78	0·262	0·85
12	21·60	3·75	0·416	6·76	0·400	6·56
24					0·085	10·50
26	9·71	5·55	0·097	12·05		

7. Comparison of the astronomical and meteorological data

The m_3 and ψ_3 in Table 3 can be directly compared. In all cases the amplitude of the ψ_3 is smaller than that of the m_3 and this is due to the neglect of the high latitude wind data. The difference for the annual term is only 4 per cent whereas that of the semi-annual term is 24 per cent. As remarked earlier the annual term reaches a maximum at mid-latitudes whereas the semi-annual term reaches a maximum at lower latitudes. Thus we would expect the discrepancy to be larger for the annual term. However, the annual wind cycle appears to be more symmetrical about the equator than the semi-annual wind cycle so that whereas the effect of the high latitude winds on the Earth's rotation largely cancels out for the annual term, this does not happen for the semi-annual term. The amplitudes of the biennial m_3 and ψ_3 agree to within 8 per cent. The agreements in phase for all three terms is excellent, particularly if the small difference in period for the biennial term is taken into consideration. That is, the biennial ψ_3 reaches a maximum for a phase angle of 158° and m_3 reaches a maximum for a phase angle of 163°. The difference between the 26-month period found for the m_3 and the 24-month period found for the ψ_3 is probably not significant. Fig. 4 shows a more direct comparison between the mean observed seasonal variations in τ for the interval 1957–1962 and the equivalent τ computed from the ψ_3. Of particular interest is the way in which the somewhat anomalous behaviour of the TU data in 1959 is identically reproduced by the wind function; the smaller than normal minimum in the spring of 1959 and the subsequent large positive value in September being due to the interference of the biennial term with the other two seasonal components.

This comparison does need further elaboration since according to Newell & Wallace (1966) the zonal wind data has a distinctly biennial component up to about the middle of 1963. The astronomical data, however, do not reveal the biennial term of the interval from April 1960 to January 1962 but shows that the term is

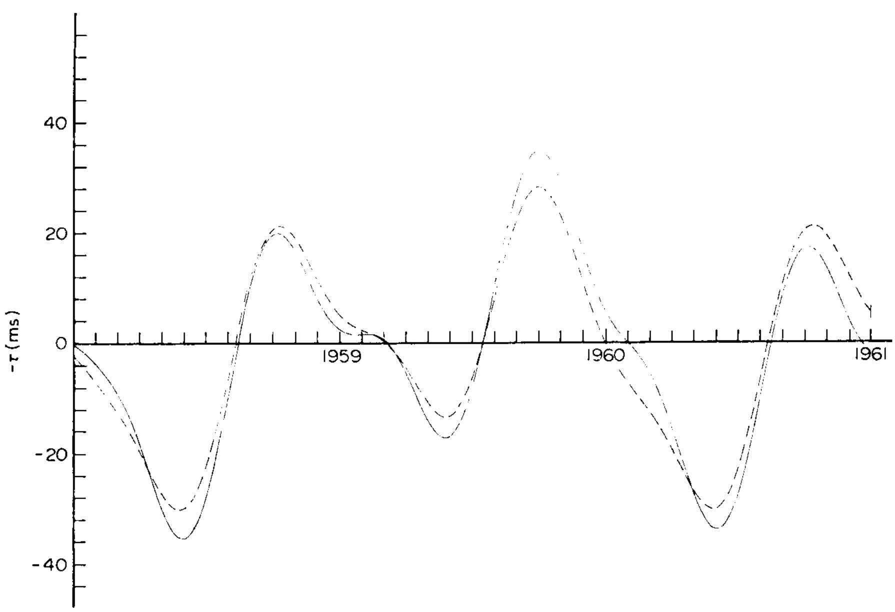

FIG. 4. Comparison of the mean variation in the observed $-\tau$ (solid line) for the period 1957–1963 and the variations computed from the mean zonal wind circulation (broken line) for the same period. The τ include the semi-annual, annual and biennial components.

reactivated during early 1962 and that it continued to exist for about a year after it has disappeared from the wind data. A tentative explanation is that for the years 1960–1961, the maximum biennial winds did not propagate downwards to their usual level. The broad global biennial cycle consists of a maximum component that propagates from about 40- or 50-km altitude to about 18 or 20 km during an interval of about 2 years. This is clearly shown in the equatorial wind profiles given by Wallace (1966), or in the profiles of Angell & Korshover (1970). When the winds reach an altitude of about 20 km the next cycle has already commenced at higher altitudes. A weak biennial oscillation however, exists down to the Earth's surface (Newell *et al.* 1972) and because of the density in the angular momentum integral (3), it is the lower atmospheric circulation that is most important as far as the Earth's rotation is concerned. Thus if the downward propagation is less pronounced than usual, the Earth's rotation is less affected. Similarly, if the cycle is not reactivated at high altitudes, the period remains evident in the Earth's rotation for some time because the previous cycle still persists—even though only weakly—in the lower atmosphere.

The Earth's rotation then suggests the following global pattern for the biennial oscillation. Prior to 1960 the biennial wind pattern was well established and propagated downwards regularly to altitudes of perhaps as much as 12–14 km to contribute sufficiently to the excitation function to explain the observed oscillation in m_3. After 1960, the cycle continued but did not significantly propagate below about 20 km and consequently the subsequent cycles did not contribute to the m_3. But by mid-1962 the maximum easterly winds propagated down to their previously low level and the cycle is re-established in the m_3 up to the middle of 1963. At higher altitudes the cycle did not continue after the middle of 1963 and consequently there is no biennial term in m_3 about 18 months later when this 'missing' cycle would have descended to a level where it would have contributed significantly to m_3. The equatorial wind profiles of Wallace (1966) for the altitudes between 16 and 32 km indicate that the westerly wind cycle in 1960–61 had a smaller amplitude than did the

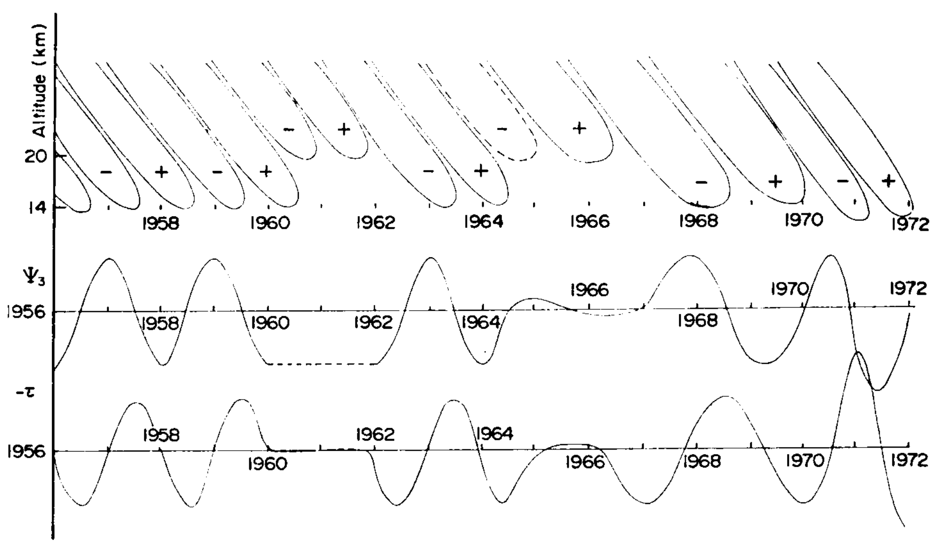

FIG. 5. Schematic representation of the zonal biennial wind circulation near the equator as deduced from the astronomical data. The lower curve represents the $-\tau$ which is an idealized representation of the observed variations shown in the lower part of Fig. 2. The middle curve is the excitation function corresponding to this $-\tau$ and the upper curve is the deduced zonal biennial wind pattern. The positive cycles indicate eastward flow.

cycles before and after this period and we would not expect this cycle to have continued downwards as far as usual. The schematic circulation pattern deduced from the Earth's rotation is illustrated in Fig. 5. We stress that this is a global pattern only but it does satisfy the astronomical data and it does not conflict with the meteorological conclusion based largely on the higher altitude winds.

For 1965 the m_3 indicates that the biennial wind, if it existed at all, did not descend below about 20–22 km but that by the latter half of 1966 a new cycle commenced at high altitudes and which appeared in the m_3 data in the spring of 1967. The period in m_3 data is not however 24 months but close to 3 years. Probably the most satisfactory explanation is that the 24-month wind cycle has not yet fully established itself and that it is variable in both phase and the altitudes at which the maximum winds occur. As m_3 is the integral of the total atmospheric circulation these variations could result in an elongation of the period of m_3. From early 1970 the biennial term in m_3 is well established and the maximum winds appear to have propagated down to altitudes of perhaps as low as 12 km.

Some of the above conclusions will be further examined when a month by month computation of the excitation function as a function of altitude is completed.

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